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THE CONTRIBUTION OF INFRARED COOLING TO THE VERTICAL MOTION FIELD AND ITS IMPLICATION IN ATMOSPHERIC ENERGETICS

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ABSTRACT

The validity of the adiabatic assumption used in estimating vertical motion is examined by comparing the relative contributions due to adiabatic processes with the diabatic process of infrared cooling. Radiometersonde data are used to prepare vertical profiles of the adiabatic and infrared components of the vertical motion. These data are first filtered to reduce the effect of random errors. Although the adiabatic component of the vertical motion is usually much larger than the infrared component, the profiles indicate that the infrared component can be important in determining the total vertical motion. A comparison of the contribution of the infrared component in cloudy versus clear sky conditions shows that this component contributes more to downward vertical motion in clear air than in the cloudy situations. The consequences of this systematic variation in estimating energy conversions is discussed. In view of these results the effect of other diabatic processes is very briefly considered.

1. INTRODUCTION

In many studies the so-called adiabatic method, which involves the first law of thermodynamics and the adiabatic assumption, is used to obtain the field of vertical motion. Despite the necessity of the adiabatic assumption, this procedure for computing vertical motions (Haltiner and Martin, 1957) is widely used in meteorology, primarily because other techniques, such as the kinematic and vorticity methods, require assumptions possibly as restrictive as the adiabatic assumption. In a recent study, Hansen and Thompson (1965) used TIROS cloud photographs and the vertical motion field obtained from both the kinematic and adiabatic methods to examine smaller scale variations in the vertical motion field. Although in general they obtained their best results using the kinematic method, they also concluded that when studying larger scale motions where datal imitations may exist, the adiabatic velocities are of more value.

In several atmospheric energetics studies the adiabatic method has been used to estimate the conversion of available potential energy to kinetic energy (White and Nolan, 1960; Jensen, 1961)—a process that involves the sinking of relatively cold air and the rising of warm air (Lorenz, 1955). The importance of using accurate vertical velocity estimates in these conversion computations cannot be overemphasized, since the magnitude of kinetic energy production determines the intensity of the general circulation. Furthermore, an understanding of the mode of the atmosphere's circulation requires a determination of the scales and mechanism by which this conversion occurs. Wiin-Nielsen (1964) notes that use of adiabatic vertical velocities to obtain the conversion of available potential energy to kinetic energy can be misleading.

In any study employing adiabatic vertical velocities, the validity of the results depends critically upon the adiabatic assumption. For certain cases high in the atmosphere, this assumption is not unrealistic. However, in other atmospheric regions, especially the lower and midtroposphere, diabatic processes of absorption and emission of radiant energy, release of latent heat, and sensible heat addition at the earth's surface tend to make estimates of the vertical motion field based on the adiabatic method less reliable. While latent heat release and sensible heat transfer are difficult to measure, radiational processes can be measured by devices such as the Suomi-Kuhn radiometersonde and satellites.

In view of the importance of an accurate determination for both energy studies and the possibility of inferring the large-scale vertical motion field from satellite data, the significance of one diabatic component—infrared radiation—upon adiabatic estimates of the field of vertical motion is investigated in this paper. Soundings of infrared cooling made at Washington, D.C., Montgomery, Ala., Green Bay, Wis., International Falls, Minn., and Amarillo, Tex., selected from the periods Dec. 20–28, 1960, and Jan. 7–18, 1961, are used to estimate the contribution of this diabatic process to the field of vertical motion. Separate vertical profiles of vertical motion associated with adiabatic processes and with the infrared component of the diabatic process are compared by using 1) a limited sample of polynomial filtered profiles and 2) a larger but unfiltered sample of profiles. Because of the importance

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of vertical motion estimates in obtaining energy conversions, the effect of neglecting the diabatic processes is discussed in some detail.

2. BASIC EQUATIONS

SYMBOLS

dh/dt=rate of heat addition per unit mass
c_p=specific heat at constant pressure
α=specific volume
R_d=gas constant for dry air
p=pressure
θ=potential temperature
$\omega=dp/dt$=vertical velocity in pressure coordinates
$\mathbf{V}_h$=horizontal wind velocity
g=acceleration of gravity
F_n=net infrared radiation
f=Coriolis parameter
u=eastward component of the wind
v=northward component of the wind

Using the notation from the list of symbols, the first law of thermodynamics may be expressed as

$$\frac{dh}{dt}=c_p\frac{dT}{dt}-\alpha\omega. \tag{1}$$

After expanding the total temperature derivative, the solution for the vertical velocity is

$$\omega=\frac{\frac{\partial T}{\partial t}+\mathbf{V}\cdot\nabla_p T-\frac{1}{c_p}\frac{dh}{dt}}{\frac{\alpha}{c_p}-\frac{\partial T}{\partial p}}. \tag{2}$$

Our discussion of vertical motion in this paper is based on the definition for vertical velocity in constant pressure coordinates and should not be confused with the vertical velocity, w, for Cartesian coordinates.

Using the Poisson equation and the equation of state, equation (2) becomes

$$\omega=\frac{-\left(\frac{\partial T}{\partial t}+\mathbf{V}\cdot\nabla_p T-\frac{1}{c_p}\frac{dh}{dt}\right)}{\left(\frac{p}{1000}\right)^{R_d/c_p}\frac{\partial\theta}{\partial p}}. \tag{3}$$

Equation (3) represents the true vertical motion, and all adiabatic and diabatic processes must be known to obtain its value. For purposes of discussion the vertical motion is divided into two components, one associated with adiabatic processes and the other with diabatic processes. The adiabatic portion of the vertical motion obtained by assuming $dh/dt=0$ is

$$\omega_A=\frac{-\left(\frac{\partial T}{\partial t}+\mathbf{V}\cdot\nabla_p T\right)}{\left(\frac{p}{1000}\right)^{R_d/c_p}\frac{\partial\theta}{\partial p}}. \tag{4}$$

The diabatic component of the vertical motion obtained by subtracting equation (4) from equation (3) is

$$\omega_D=\frac{\frac{1}{c_p}\frac{dh}{dt}}{\left(\frac{p}{1000}\right)^{R_d/c_p}\frac{\partial\theta}{\partial p}}. \tag{5}$$

Equation (4) is the usual expression for the adiabatic component of the vertical motion field, and equation (5), which represents the diabatic effect, will for convenience be designated the diabatic component of the vertical motion field. In this study, only nonadiabatic heating due to the field of terrestrial radiation is considered. Under steady-state conditions and horizontal isotropy for the field of infrared irradiance, the heat addition per unit mass due to infrared divergence is

$$\left(\frac{dh}{dt}\right)_I=g\frac{\partial F_n}{\partial p}. \tag{6}$$

Thus the diabatic component of vertical motion due to the field of infrared irradiance is

$$\omega_I=\frac{\frac{g}{c_p}\frac{\partial F_n}{\partial p}}{\left(\frac{p}{1000}\right)^{R_d/c_p}\frac{\partial\theta}{\partial p}}. \tag{7}$$

The adiabatic estimate of the vertical motion field is obtained using (4), and the infrared diabatic component is estimated by (7) from radiometersonde data (Suomi-Kuhn, 1958). The relative importance of *infrared processes* is examined by comparing profiles of this *diabatic* component of the vertical motion field with profiles using the *adiabatic* assumption.

3. COMPUTATIONAL PROCEDURES

ADIABATIC ESTIMATES OF VERTICAL VELOCITIES

The calculation of adiabatic vertical velocities by equation (4) requires estimates of the local temperature tendency, horizontal temperature advection, and the stability measure. An inherent problem in using synoptic data is that the local temperature tendency is a time-averaged determination while the horizontal advection by the geostrophic wind field is a spatially determined quantity at a given time (Panofsky, 1951). The infrared component is calculated from data taken at the same time as that used in computing the advection term. In order to obtain diabatic and adiabatic profiles that are representative for the same time, the local temperature tendency has been determined from observations taken 24-hr apart while the advection term of the adiabatic equation and infrared component computation were made from data taken at the midpoint of the 24-hr period. For compatibility we assume that the 24-hr temperature tendency is representative and is the best estimate corresponding to the radiometersonde data.

The above assumption seems reasonable because the free-atmosphere temperature time series tends to be a smooth function whose principal variation is associated with large scales of motion with intervals of several days between adjacent relative temperature maxima and minima. For this scale of variation, a second order Taylor's series expansion about the midpoint time t_0 should provide a valid description of the temperature in the 24-hr interval from t_0-12 hr to t_0+12 hr. If this condition is satisfied, then the finite estimate of the 24-hr time-average temperature tendency is a valid estimate for the local tendency at t_0.

In this study, the horizontal temperature advection is determined through the thermal wind relationship. This allows the advection computation to be made solely from the wind data for a single station and thus avoids the problem of obtaining a geostrophic wind determination from the large-scale pressure pattern. Under the assumption that the acceleration and frictional force are constant with height, the vertical shear of the horizontal wind is

$$\frac{\partial \mathbf{V}_h}{\partial p}=-\frac{R_d}{fp}(\mathbf{k}\times\nabla_p T).$$

A rearrangement and scalar multiplication by $\mathbf{V}_h$ yields the horizontal temperature advection

$$\mathbf{V}_h\cdot\nabla_p T=\frac{fp}{R_d}\left(v\frac{\partial u}{\partial p}-u\frac{\partial v}{\partial p}\right). \tag{8}$$

The accuracy of the temperature advection computed with equation (8) depends upon the method used to evaluate u, v, $\partial u/\partial p$, and $\partial v/\partial p$. Since the u- and v-components contain random errors plus small-scale wind variations, serious errors in the estimation of these quantities result if a method is not used to reduce the influence of random errors and small-scale variations, both of which are essentially noise superimposed upon the synoptic-scale variations.

One way of reducing the effect of noise is to use least squares approximating polynomials for filtering. Approximating polynomials (Hildebrand, 1956) can be used whenever the basic observations are discrete measurements of a smooth true function. The advantage of using least squares approximating polynomials is that an exact fit to the basic data is avoided. An exact fit requires the estimated function to pass through each data point, thus oscillating about the true function. On the other hand, approximating polynomials supply a better estimate of the smooth true function by suppressing the effects of random errors and small scale wind variations.

In a portion of this study the u- and v-component profiles are filtered by fitting a second-degree polynomial to five adjacent observations that are vertically spaced at 50-mb intervals. Pressure is used as the independent variable for filtering. The resulting polynomials, of the form $\hat{u}=u(p)$ and $\hat{v}=v(p)$, are then differentiated, and a set of filtered estimates $\partial\hat{u}/\partial p$ and $\partial\hat{v}/\partial p$ are obtained at the midpoint of five observations. The next set of filtered estimates is made by adding a new observation adjacent to the upper pressure level and deleting the observation at the lower pressure level and then predicting the new mid-observation. After obtaining values of $\hat{u}$, $\hat{v}$, $\partial\hat{u}/\partial p$, and $\partial\hat{v}/\partial p$ from 850 mb to 200 mb at 50-mb increments, filtered estimates of the horizontal temperature advection are computed.

Figure 1, which presents profiles of the raw and filtered values of the u-component at Washington, D.C., at 00 GMT on Dec. 22, 1960, clearly illustrates the filtering effect of the least squares approximating polynomials. In figure 2 the estimates of $\partial u/\partial p$ computed by a finite-difference approximation given by

$$\left(\frac{\partial u}{\partial p}\right)_p\simeq\frac{U_{\left(p+\frac{\Delta p}{2}\right)}-U_{\left(p-\frac{\Delta p}{2}\right)}}{\Delta p}$$

for two different pressure intervals ($\Delta p=100$ mb and 200 mb) and those obtained by differentiating the second-degree approximating polynomial estimate of the u-component profile are compared. All finite differences were computed using the basic observations. The 100-mb finite-difference estimates include a random oscillation that does not accurately characterize the behavior of the large-scale wind profile. Both the 200-mb finite difference and the polynomial technique provide smoother and better behaved estimates at each level, with the polynomial

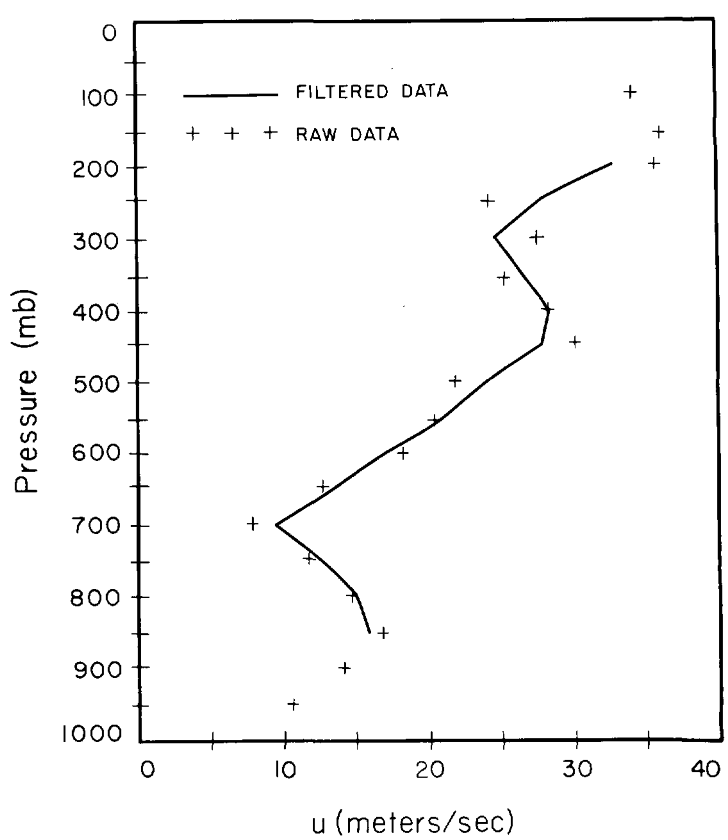

FIGURE 1.—Comparison of raw and filtered values of the eastward component of the wind at Washington, D.C., Dec. 22, 1960, 00 GMT.

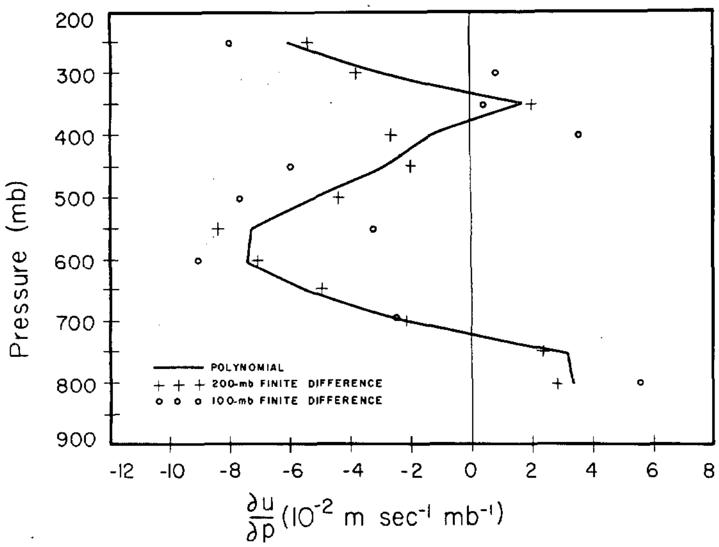

FIGURE 2.—Comparison of filtered and finite-difference estimates of wind shear for the profile in figure 1.

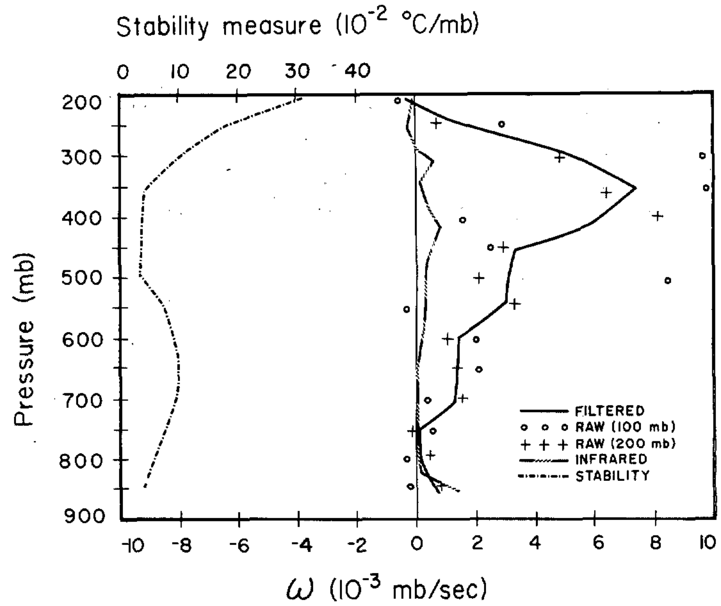

FIGURE 3.—Clear sky profiles of adiabatic and infrared components of vertical motion at Washington, D.C., on Dec. 20, 1960, 00 GMT.

giving the somewhat smoother profile. These two estimates agree closely due to similar weighting in the computations. The smoother appearance of the polynomial estimate can probably be attributed to the fact that the random error variance of the polynomial estimate is 80 percent of the random error variance of the finite difference. This can be verified from an error analysis of the two methods. Furthermore, by Gauss's theorem, differentiated polynomial estimates are minimum variance estimates if polynomial estimates are unbiased (Johnson, 1965).

Estimates of $\partial u/\partial p$ and $\partial v/\partial p$ were also made with a finite-difference approximation with $\Delta p=300$ mb and also with a polynomial fit to seven data points, but in some cases this procedure tended to remove significant features of the wind profile. For this reason, the polynomial fit to five adjacent observations to determine $\hat{u}$, $\hat{v}$, $\partial\hat{u}/\partial p$, and $\partial\hat{v}/\partial p$ was selected for computing horizontal temperature advection. The nonfiltered advection estimates, which are presented for comparison, were determined from observed wind component values and a 200-mb finite evaluation.

The stability measure was also computed by differentiating the second-degree polynomial fit of potential temperature with respect to pressure and, for comparison, by a 200-mb finite-difference approximation. Values obtained by these methods agree very well with few exceptions. In this sample the improvement due to filtering the stability measure was not as striking as anticipated because the static stability of the data for the middle-latitude winter troposphere is large. In tropical regions and summer conditions where the tropospheric lapse rate approaches the dry adiabatic, filtered estimates of stability would provide a significant improvement. Included in figures 3, 4, 5, and 6, which are discussed in section 4, are filtered profiles of the stability measure. Because of the abrupt change in the temperature and potential temperature profiles at the tropopause level, the stability measure estimates should not be computed in the same way they are at levels above and below the tropopause. If the conventional five-point filtering or finite-difference technique with $\Delta p=200$ mb is used to evaluate stability at or within 100 mb above or below the tropopause, the sudden change in stability that actually occurs is partially removed by smoothing. For this reason, values of static stability determined from the highest complete 200-mb interval in the upper troposphere should be used for those levels below the tropopause that otherwise could not be computed without using lower stratospheric data. A similar procedure for determining stratospheric static stability should be applied in the region just above the tropopause. Since we have not incorporated this procedure in our present study, care must be exercised in the comparison of magnitudes of the two biased estimates in the vicinity of the tropopause.

INFRARED COMPONENT OF THE VERTICAL VELOCITIES

The infrared cooling contribution to the field of vertical motion was evaluated by applying equation (6) to data obtained from radiometersonde flights made at 00 GMT at Washington, D.C., Montgomery, Ala., Green Bay, Wis., International Falls, Minn., and Amarillo, Tex., during the periods of Dec. 20–28, 1960, and Jan. 7–18, 1961. Diabatic vertical velocity profiles were obtained by differentiating the approximating polynomial description of the profiles of net irradiance and potential temperature in a manner similar to that used to obtain the filtered wind estimate. However, for infrared component profiles the independent variable for filtering is time rather than

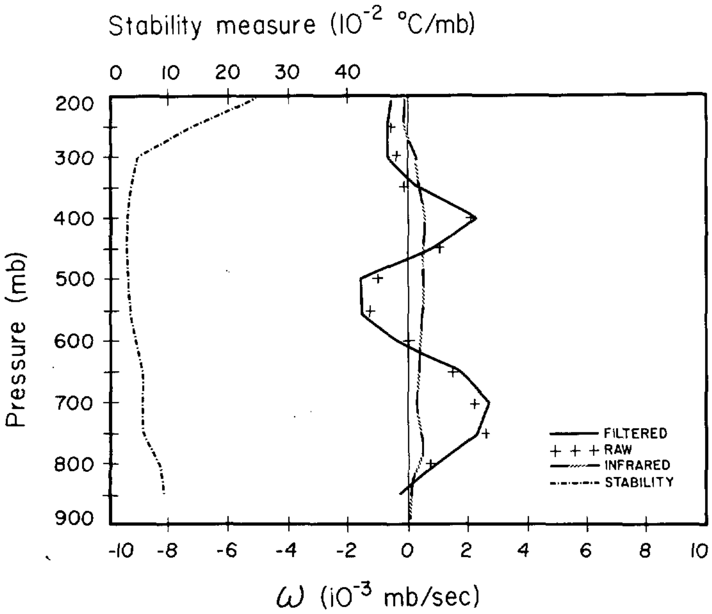

FIGURE 4.—Clear sky profiles of adiabatic and infrared components of vertical motion at Washington, D.C., on Dec. 26, 1960, 00 GMT.

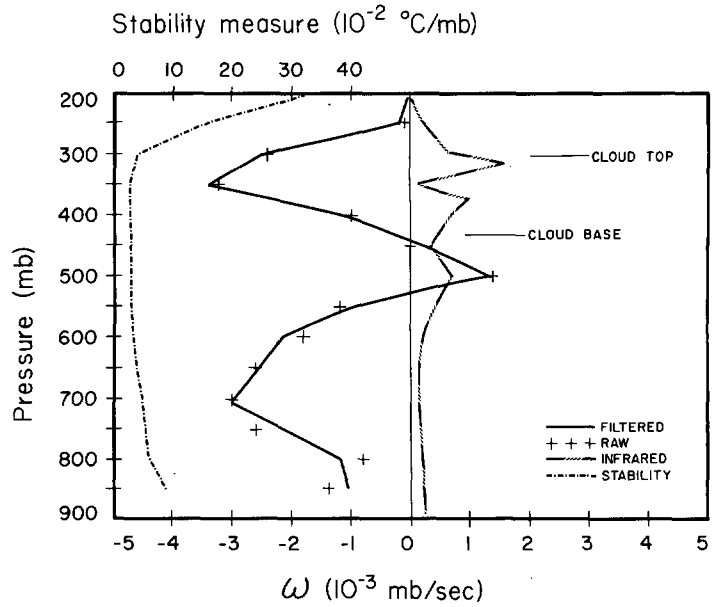

FIGURE 5.—Cloudy sky profiles of adiabatic and infrared components of vertical motion at Green Bay, Wis., on Jan. 17, 1961, 00 GMT. Cloud cover was reported as 0.8 cirrostratus.

pressure. This technique of polynomial filtering of radiometersonde data has been presented by Kuhn and Johnson (1966).

4. DISCUSSION OF RESULTS

In this study the effect of neglecting infrared cooling when computing adiabatic vertical motions is considered by 1) comparing individual profiles of the adiabatic component obtained from filtered data with the infrared component for clear and cloudy regimes, 2) employing a synoptic case study of the time variation of the adiabatic and infrared profiles, and 3) comparing mean profiles of the infrared with the adiabatic component obtained from a relatively large sample of unfiltered data, also for clear and cloudy conditions.

Before comparing the adiabatic and diabatic components, the effect of finite-difference and filtering techniques in estimating the adiabatic vertical velocities in a clear atmosphere are presented in figure 3. Profiles of adiabatic vertical velocity obtained by use of filtered estimates of the horizontal temperature advection and stability are shown by the continuous lines. The effect of the pressure interval used in finite-difference evaluations of the temperature advection and stability measure for estimates of the adiabatic vertical velocities is shown for two cases, $\Delta p=100$ mb and $\Delta p=200$ mb. The 100-mb-interval estimates portray a profile of adiabatic vertical motion that is not characteristic of large-scale atmospheric motions. The extreme oscillations very likely reflect the presence of small-scale variations and random errors in wind and temperature data. If a 50-mb interval is used, we would expect the resulting profiles to exhibit an even more pronounced oscillatory behavior.

Comparison of the 200-mb nonfiltered and filtered estimates reveals good agreement; however, in this example, the filtered profile is somewhat smoother than the nonfiltered one, and the region of maximum adiabatic vertical motion is displaced upwards by about 50 mb.

INDIVIDUAL CLEAR AND CLOUDY SKY PROFILES OF ADIABATIC AND INFRARED COMPONENTS OF VERTICAL MOTION

In addition to adiabatic profiles of vertical motion, figure 3 also presents the infrared component profile for this clear sky case. In this instance, the 200-mb filtered and nonfiltered estimates of the adiabatic vertical velocities agree well with our concept of vertical motion in a clear atmosphere, with positive values of ω_A throughout most of the troposphere. Likewise, the infrared component, though small, is predominantly positive.

Unfortunately, not all vertical motion profiles from clear sky cases are as consistent as that presented in figure 3. For example, figure 4 shows a reasonable infrared component profile, but an adiabatic profile that displays negative values of ω_A through a portion of the middle troposphere. The fact that no cloud cover was observed in this case could be attributed to a number of reasons. Cloud formation is not instantaneous but requires some finite time depending upon the initial degree of saturation. In support of this reasoning is the observation by Hansen and Thompson (1965) of a time lag between the initiation of upward vertical motion determined kinematically and cloud formation as observed by TIROS photographs. In some situations, but not necessarily this particular clear

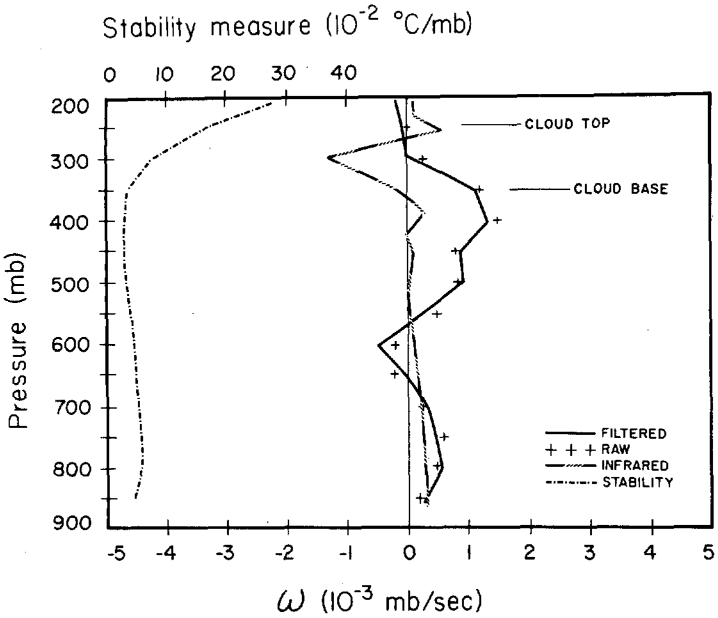

FIGURE 6.—Cloudy sky profiles of adiabatic and infrared components of vertical motion at Montgomery, Ala., on Jan. 12, 1961, 00 GMT. Cloud cover was a cirrostratus overcast.

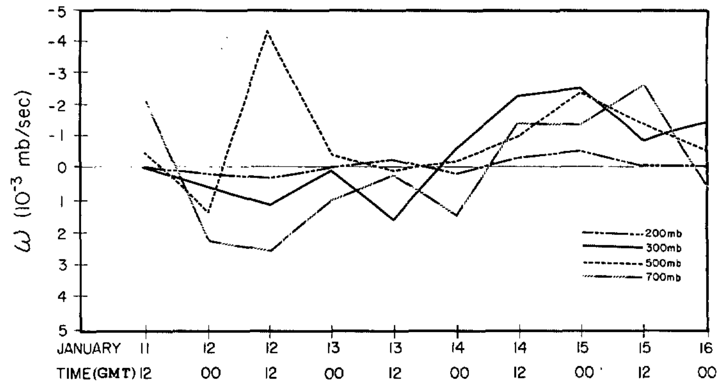

FIGURE 7.—Time cross-section of the adiabatic component of vertical motion at four selected levels at Washington, D.C., for the period of Jan. 11–16, 1961.

case, the neglect of other diabatic effects leads to erroneous estimates of vertical motion. Finally, the possibility exists that errors were present in the data.

The infrared component was largest, as expected, in the presence of clouds. Figure 5 shows the profiles for Green Bay, Wis., on an evening with 0.8 cirrostratus overcast. In this case, the adiabatic vertical velocity maximum near 350 mb corresponds closely to the cloud level while the maximum infrared component occurred in the vicinity of the cloud top. The cloud top coincides very closely with the tropopause and the sudden decrease in adiabatic vertical velocities in the region above.

A second example of vertical velocity profiles for a cirrostratus overcast condition is presented in figure 6. In this case, the adiabatic contribution to the total vertical motion is positive almost everywhere below the cloud, which appears inconsistent with the maintenance of a cloud.

CASE STUDY

To illustrate temporal variations that occurred in the components of vertical motion with changes in the synoptic weather pattern, the 4-day period of Jan. 12–15, 1961, at Washington, D.C., is considered. During the first 2 days, skies at Washington were clear as high pressure dominated the eastern United States. However, on January 13, a low-pressure area that had formed off the Texas Gulf Coast was just south of New Orleans and moving northeast at about 15 kt. There was a cold-core Low aloft to 300 mb over central Texas to the west of the surface depression. By 06 GMT on January 14, the low-pressure center was located just west of Tallahassee, Fla., with a cold front along the western Florida coast and a warm front through southern Georgia into the Atlantic. At 00 GMT on January 14 there was scattered cirrus, and at 06 GMT scattered stratocumulus was present over Washington. On January 15 there was a nimbostratus overcast with continuous rain as the low-pressure center moved southeast of Washington on a northeasterly course.

Figure 7 shows the time variation of the adiabatic vertical velocities at four selected pressure levels. During the first 2 days of the period, the adiabatic method indicated general subsidence at all levels except 500 mb. The large negative value of ω_A on January 12 probably results from an observational error. By 12 GMT on January 14 the adiabatic velocities indicated generally upward motion, and by the end of the fourth day the strength of the upward motion was decreasing at most levels.

Figure 8 presents the infrared component profiles at 00 GMT for each day of the period. The first 2 days show profiles typical of clear skies, while the third day shows some effect of cirrus with bases located at about 250 mb. Examination of the profile for January 15 shows a more pronounced effect with the top of the highest cloud layer at about 300 mb. Apparently the effective emissivity of the cirrus layer on January 15 was much greater than on January 14. The lack of variation in the profile below 550 mb indicates nearly radiational equilibrium below this level.

Adiabatic vertical-velocity estimates in this case study showed generally good agreement with our concept of the vertical motion field for the synoptic pattern, a result that was not true in all cases. Also, the diabatic component corresponded well with cloud conditions and demonstrated its importance in the cloudy regions.

MEAN CLEAR AND CLOUDY PROFILES

Although filtered profiles of adiabatic velocities were used for comparison with the infrared components in individual cases, it was not possible, because of the limited number of cases, to obtain mean profiles that were significant. The limited number of filtered adiabatic

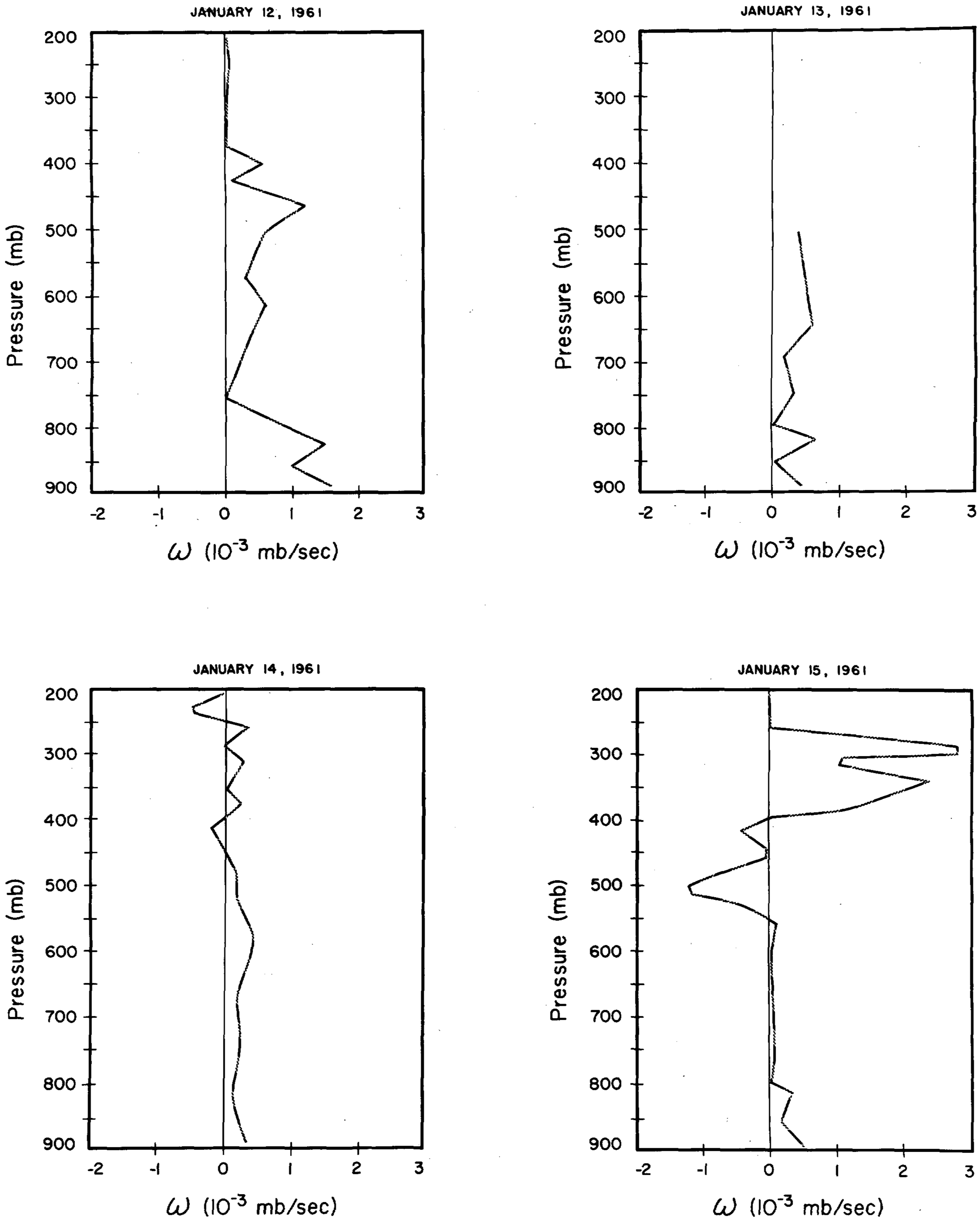

FIGURE 8.—Vertical profiles of the infrared component of vertical motion at 00 GMT for 4 days of the case study.

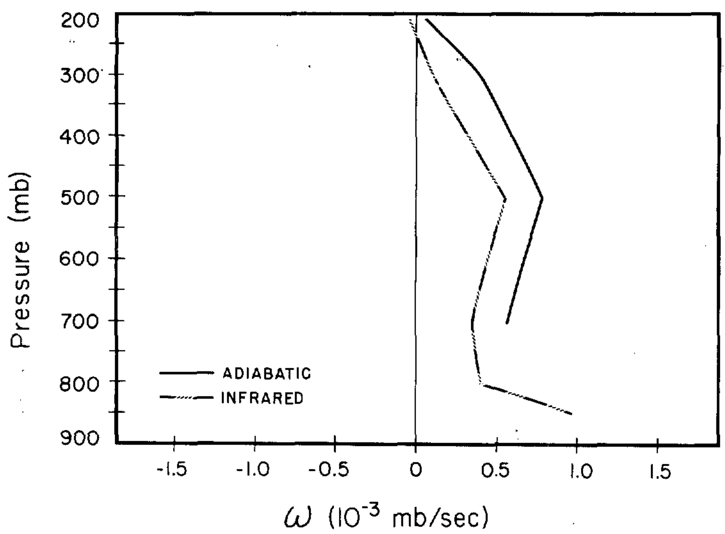

FIGURE 9.—Mean profiles of the adiabatic and infrared components of vertical motion for clear sky conditions, compiled from 21 cases.

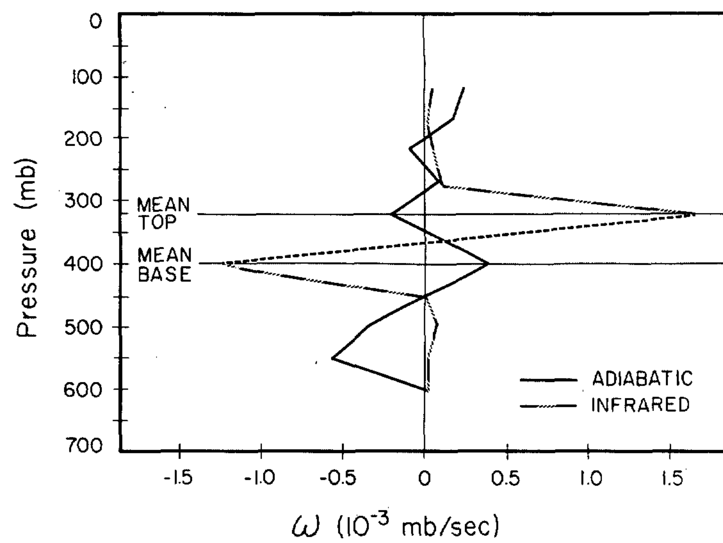

FIGURE 10.—Mean profiles of the adiabatic and infrared components of vertical motion for cloudy sky conditions. Profiles are constructed with respect to the mean cloud position of the 10 cases used.

profiles resulted from the fact that data at 50-mb increments, which were necessary for application of the filtering technique, were available for only some cases. Consequently, mean profiles of infrared and unfiltered adiabatic velocities were prepared from a relatively large sample of data obtained at standard levels.

Mean profiles of adiabatic vertical velocities and the corresponding infrared components were obtained from 31 individual profiles computed from observations taken during the data period described in the Introduction. As in the case of filtered profiles, equations (4) and (7) were used to compute the adiabatic and the infrared components, respectively. The infrared data used to estimate mean profiles were computed in the same way as in the individual profiles. However, the unfiltered adiabatic component used to obtain mean profiles differed from the individual filtered profiles in that the time for which they are valid differs by 6 hr from the time for which the infrared profiles were calculated. This time difference resulted from the fact that the local temperature tendency was computed from observations taken 12 hr rather than 24 hr apart.

Separate mean profiles were prepared for clear and cloudy situations. To be classified as cloudy, at least 0.8 coverage of middle or high clouds was required; and to be classified as clear, less than 0.1 cloud coverage was necessary. Figure 9 presents mean profiles of adiabatic and infrared components computed using 21 clear cases. Both profiles correspond with the concept of subsidence in the presence of clear skies. The mean adiabatic profile shows positive values of ω_A throughout, averaging about 0.4×10^{-3} mb sec^{-1}, with a maximum of 0.75×10^{-3} mb sec^{-1} at 500 mb. The infrared component averages about 0.25×10^{-3} mb sec^{-1}. The small mean for the adiabatic component indicates considerable heterogeneity in individual profiles for clear conditions. In contrast, the relatively large magnitude of the mean for the infrared component indicates less heterogeneity for this class. The slight negative value at 200 mb is associated with convergence of infrared irradiance in the region of the tropopause.

Figure 10 presents mean profiles obtained from 10 cloudy cases. In preparing this figure, a technique used by Kuhn (1966) was employed to determine the pressures at the base and top of the cloud layer for the individual profiles of net irradiance. Once these pressures were established for each case, infrared components were computed at 50-mb intervals to a maximum of 200 mb above the cloud top and 200 mb below the cloud base. These values were then averaged to obtain the mean profile. The adiabatic contribution at these levels was obtained by interpolating between the values that had been computed at standard levels for the individual cases and then averaging those values. The dotted line within the cloud connecting the infrared components at the base and top should not be considered as representing the infrared component within the cloud, since most of the cloud interior is in radiation equilibrium.

It should be noted that the relatively large negative value of ω_I at the mean cloud base and large positive value at the top are mainly due to a net absorption of infrared radiation at the base and emission at the top and are not representative of large-scale vertical motion. However, these radiation processes are important in enhancing small-scale turbulent motion in clouds. The heating at the base and cooling at the top decreases the static stability within the cloud and consequently may lead to turbulent mixing and an increased upward heat transfer. Möller (1951) estimated that, in a cloud 0.5 km thick at an elevation of 2 km, only 45 min would be required to increase the lapse rate from isothermal to 0.5°C/100 m, while in a cloud of the same thickness at an elevation of 5 km only 20 min would be needed to produce the same change. Möller assumed that no convective transfer of heat occurred during the destabilization process.

Although infrared processes are important in clouds, the contribution of latent heat to vertical motions can be more significant. Immediately beneath a precipitating cloud layer the evaporating droplets cause cooling, which may offset any infrared warming (and negative values of ω_I), but at lower levels the evaporation may reinforce any small positive values of ω_I. In the cloud, latent heat release due to condensation gives a negative contribution to the total vertical motion. At the cloud top there can be both condensation and evaporation of moisture. The condensation is the result of ascending motion, while evaporation results because of turbulent mixing and subsequent entrainment of drier air from above. Whether heating or cooling is occurring at the cloud top depends upon which process is predominant.

Although the diabatic component of vertical motion in the vicinity of cloud top and base must be neglected when considering the large-scale vertical motion field, an examination of figures 9 and 10 shows an important difference in the diabatic profiles that are representative of large-scale atmospheric processes. Infrared cooling beneath an extensive cloud layer is typically less than in clear conditions. For example, the average value of the infrared component of vertical motion in the mean cloudy profile (fig. 10) between 450 and 600 mb is very close to zero, while the average value of this component in the mean clear profile (fig. 9) in the same pressure interval is approximately 0.4×10^{-3} mb/sec.

5. CONSEQUENCE IN ENERGY CONVERSION ESTIMATES

Although the primary purpose of our study is to compare the infrared with the adiabatic component of vertical motion, it is also important to consider the significance of these results in atmospheric energy studies. In this study the contribution of infrared cooling to the total vertical motion field was usually positive, but much more so in clear areas than in regions with middle or high clouds. Since growing and organized weather disturbances are now recognized as major sites of conversion of potential to kinetic energy in middle latitudes (Starr, 1958), it is important to consider the magnitude of the bias error resulting from the use of adiabatic velocities (Wiin-Nielsen, 1964) in estimating the production of kinetic energy by the $\omega\alpha$ integral. In troughs associated with growing and organized disturbances, cloudy areas tend to be associated with the warmer air to the east and clear areas with the colder air in and below the midtropospheric trough. Since the infrared component of vertical motion tends to be in phase with the adiabatic component, the actual conversion will tend to be underestimated if the infrared component is neglected. Our results can be used to estimate that portion of the bias error due to the neglect of diabatic processes.

The general expression for the conversion of eddy available potential energy to eddy kinetic energy is

$$C(P,K)=-\int \omega'\alpha' dm \tag{9}$$

where m is the mass, the primes denote deviations from a latitudinal mean on a pressure surface, and the integration extends over the entire atmosphere. The true but unknown eddy conversion, indicated by asterisks, is

$$C^*(P,K)=-\int \omega'^*\alpha'^* dm. \tag{10}$$

Since the total vertical motion field is given by the sum of adiabatic and diabatic portions, equation (10) may be written as

$$C^*(P,K)=-\int (\omega'^*_A+\omega'^*_D)\alpha'^* dm. \tag{11}$$

If we assume that estimates of specific volume and adiabatic vertical velocity are unbiased and their random observational errors uncorrelated, the expected value of conversion estimates using the adiabatic assumption is

$$E[C(P,K)]=-\int \omega'^*_A\alpha'^* dm=C^*(P,K)-\delta \tag{12}$$

where

$$\delta=-\int \omega'^*_D\alpha'^* dm. \tag{13}$$

Equations (12) and (13) show that the bias error δ is given by the covariance of the diabatic component and the specific volume. The diabatic component of the vertical velocity is

$$\omega_D=\omega_S+\omega_I+\omega_L+\omega_C+\omega_F \tag{14}$$

where the subscripts denote the following diabatic components:

S=direct solar absorption,
I=emission of infrared energy,
L=release of latent heat,
C=sensible heat addition by conduction at the earth's interface,
F=internal heat addition due to frictional dissipation of kinetic energy.

In the developing and organized stages of the weather-producing disturbance, the diabatic components of the release of latent heat, infrared cooling, and solar absorption are in phase with the adiabatic vertical motion and the specific volume fields. The component of sensible heat addition at the earth's surface is opposite in phase to the field of vertical motion, since the coldest air at the surface will be heated more than the warm air. The frictional component is unimportant, for there should be little covariance between specific volume or ω and the frictional dissipation of the horizontal kinetic energy. Thus, the systematic bias error due to the neglect of the four heating components is

$$\delta=-\int(\omega'^*_S+\omega'^*_I+\omega'^*_L+\omega'^*_C)\alpha'^*dm \tag{15}$$

where the first three terms of the integral tend to be positive, causing the adiabatic conversion estimates to be low while the last term tends to be negative, causing the adiabatic conversion estimates to be high.

Our results, although tentative, allow a quantitative estimate for the infrared component. The results shown in figures 9 and 10 indicate that the infrared diabatic component may be expressed as

$$\omega'_I=b_I\omega'_A \tag{16}$$

where the value of b_I lies at least between 0.1 and 0.2 and is possibly greater. For regions of the quasi-geostrophic planetary waves (i.e., those without embedded, smaller organized disturbances) for which the adiabatic assumption is valid or for unorganized scales in which there is probably no systematic relation between the diabatic component of vertical motion and the adiabatic vertical velocities, b_I may be set to zero without inducing estimation errors in the conversion integral. Under these approximations the infrared bias error component in adiabatic vertical velocities is

$$\delta_I=-\left\{\int b_{I_1}\omega'^*_{A_1}\alpha'^*_1dm_1+\int b_{I_2}\omega'^*_{A_2}\alpha'^*_2dm_2\right\} \tag{17}$$

where the subscript one denotes values from atmospheric regions possessing organized disturbances, and the subscript two denotes values associated with quasi-geostrophic scales and unorganized circulations. If one assumes that 50 percent of the estimated conversion by the adiabatic method occurs in organized disturbances of smaller synoptic scales and 50 percent in the larger planetary scale without an organized heating field in which b_{I_2} is zero, the bias infrared error component related to the expected value of adiabatic estimates is

$$\delta_I=\frac{b_{I_1}}{2}E[C(P,K)] \qquad 0.1\leq b_{I_1}\leq 0.2. \tag{18}$$

Thus, estimates of the conversion of potential energy are underestimated, and the actual adiabatic estimates should be adjusted upward by 5 to 10 percent to remove the infrared component bias and to achieve better estimates of the true conversion.

Undoubtedly there is a similar significant and more striking bias error due to the neglect of the effects of latent heat in the precipitating regions of the organized disturbance. The results obtained by Danard (1964) for ω_A and ω_L indicate that a possible range for b_L is from 1.0 to 1.5. Thus, it is possible that actual adiabatic estimates should be adjusted upward by 50 to 75 percent to remove bias from the neglect of latent heat release.

With respect to the bias conversion error introduced by the neglect of effects of sensible heat addition at the earth's interface, the systematic bias error, δ_C, should be negative. Sensible heat addition at the earth's interface will always be larger in the colder air than in the warmer air. This is particularly true in the winter when cold continental air flows off the eastern portions of both the Asian and North American Continents. The effect of neglecting the component of the vertical motion associated with sensible heating and use of the adiabatic vertical component is to overestimate the magnitude of the downward vertical motion. In the boundary layer, the magnitude of the true vertical motion should be small; and in the presence of strong sensible heating of the cold air, the adiabatic component, which is a large positive value, may be nearly balanced by a large negative diabatic component. Thus the sign of b_C, corresponding to the definition of b_I, is probably negative, and δ_C is also negative. Although we are unable to estimate the magnitude of this bias error, we speculate that it is opposite in sign and larger in magnitude than the infrared bias error but less than the latent heat bias error. When adiabatic vertical velocities and the $\omega\alpha$ integral are used to estimate energy conversions, we would emphasize Jensen's (1961) statement ". . . that the occurrence of very intense energy transformations within the boundary region is questionable in view of certain nongeostrophic and nonadiabatic effects."

Wiin-Nielsen (1964), using Jensen's (1961) calculations of the adiabatic conversion of available potential energy to kinetic energy for individual layers, estimated that for the Northern Hemisphere the conversion due to adiabatic motions amounted to 4.24 watts m^{-2} during January 1958 and 2.71 watts m^{-2} during April 1958. Our crude estimates of the vertical motion bias errors produced by diabatic effects indicate that these energy conversion estimates should be increased. With an increase, they would compare more favorably with Kung's (1966) earlier estimate of 6.4 watts m^{-2} and his later estimate (1967) of 4.12 watts m^{-2} for frictional dissipation and Dutton and Johnson's (1967) summary of the transformation rates involved in the atmospheric energy cycle. Admittedly, our comments concerning the bias errors due to neglect of diabatic processes are quite speculative; however, our results for infrared processes, Danard's for latent heating, and the questionable effects of sensible heating indicate that additional investigations of the energy conversion processes should be undertaken. It is quite possible that the use of adiabatic vertical motions in estimating the energy conversions on a hemispheric scale produce estimates which may be significantly low.

6. CONCLUSIONS

Although the adiabatic method is widely used to estimate large-scale vertical motion and is probably as effective as any of the other commonly used methods, the results of this study indicate that its limitations are considerable. If data from small intervals of pressure are used (e.g., 50 mb) to obtain profiles of adiabatic vertical velocities, it appears that some type of filtering technique such as that used here should be employed to reduce the random errors. In general, considerable variations exist between individual adiabatic profiles, even when the data are filtered and classified according to the amount of cloudiness.

Although in individual cases the infrared contribution to total vertical motion tends to be quite small, the contribution is quite consistent from one case to another. Thus when mean profiles of adiabatic and infrared velocities are prepared, the contribution of infrared processes appears quite significant, perhaps amounting to 30 percent of the adiabatic contribution. In regions of clear skies the mean infrared contribution to the downward motion is about 60 to 70 percent of the adiabatic component. In cloudy regions the infrared contribution is practically zero below the cloud, while above the clouds it is just slightly positive. In the presence of a cloud layer the infrared radiation process undoubtedly plays an important role by reducing the hydrostatic stability within the cloud layer and thus increasing the vertical transfer of heat.

The diabatic processes also reduce the reliability of adiabatic estimates of conversion of available potential energy to kinetic energy within the cyclone scale. Since, within this scale, the diabatic components of infrared radiation, the absorption of solar energy, and release of latent heat are in phase while the component of sensible heat addition is out of phase with the adiabatic component of vertical motion, the net effect of these diabatic processes should be considered. It appears that in a disturbance the diabatic contribution to the vertical motion field may be sufficient to result in a substantially larger energy conversion than that obtained in studies based only on adiabatic motions.

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